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A STUDY OF INVERSE METHODS FOR PROCESSING OF RADAR DATA

Mohamed F. Chouikha, Ph.D.

Howard University
Department of Electrical and Computer Engineering
2400 Sixth Street NW
Washington, DC 78235

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Air Force Research Laboratory
Human Effectiveness Directorate
Information Operations and
Special Programs Division
2486 Gillingham Dr.
Brooks City-Base, TX 78235

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RICHARD A. ALBANESE
Project manager

//SIGNED//
JAMES W. RICKMAN, MAJ, USAF
Chief, Special Projects Branch

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14. ABSTRACT

This is a special report for the project "A Study of Inverse Methods for Processing of Radar Data" supported by the DoD US Air Force Research Lab. This project started on July 1, 2004 and continued until August 31, 2005. The goal of this project is to investigate the possibility of new inversion algorithms for radar image processing to improve signal quality and reduce the effects of clutter based on study of known geophysical inversion algorithms. Results of processing synthetic data shows promise to future processing of actual radar data.

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A Study of Inverse Methods for Processing of Radar Data

Mohamed F. Chouikha, Ph.D.

Summary

This is a technical report for the project "A Study of Inverse Methods for Processing of Radar Data" supported by the DoD US Air Force Research Lab. This project started on July 1, 2004 and continued until August 31, 2005. The goal of this project is to investigate the possibility of new inversion algorithms for radar image processing to improve signal quality and reduce the effects of clutter based on study of known geophysical inversion algorithms.

1. Introduction

The scattering of energy can be considered on a point by point basis, each diffraction point contributing a part of the total signal. Kirchhoff methods pioneered by Bleistein and Cohen at the Colorado School of Mines [6] used an asymptotic integral method to construct a migration image of the seismic data. By reduction for the number of integrations they were able to make this method computationally feasible. The method needs accurate travel times for the primary wave front to every subsurface point from each source receiver location. Both ray tracing and eikonal schemes have been used to computer these travel times. A by-product of their scheme is a local estimate of the parameters beneath the reflector surfaces constructed from two different integral approximations.

Claerbout [9] and his Stanford Exploration Project took the approach of downward continuing the surface data using a paraxial approximation and its extensions to larger angles. Their further work in this area has developed angle versus aperture imaging algorithms using the one way wave equation.

The use of angle dependent plane waves for seismic imaging was invented at the Amoco Research Center by Dan Whitmore. He was able to show in his PhD thesis that using a few plane wave angles and converting the time/distance data into ray parameter and time or tau-p data that very accurate images can be developed. Gazdag (1978) [7] and then Stoffa et al. (1990) [4] made use of phase shift and interpolation phase shift schemes to develop very good migration algorithms, again using the one way wave equation.

All of these methods make transformations to the wave equation. Another computer expensive but very successful method uses the approach of running the wave equation backward in time. This uses the inverse of the modeling equation for imaging. First published by Hemon [11], it was popularized by Whitmore [13] after Kelly and team [12] demonstrated successful acoustic modeling codes. Currently for real three-dimensional elastic imaging, prestack reverse time methods [10] are the most accurate and are competitive in cost.

A distinct characteristic of geological interface surfaces is the relatively long specular surfaces in relationship to the surface source/receiver geometry. To extract subsurface features like river channels and fractures from such regular structures has posed a significant challenge. Several researchers in the seismic arena have constructed coherency measures to highlight these relatively short features. The eigenvalue method of Gerztenkorn is typical of these seismic coherency methods. Within the Fresnel zone the resolution of good quality seismic data is thought to be on the order of a quarter of a wavelength. With the much higher frequencies of radar and at the resolution needed for imaging vehicles the expected number of wavelengths in a typical data volume will be much larger than in seismic. This will impose further restrictions on the use of these seismic based methods, but modeling studies will demonstrate with these algorithms their relative merits.

Signal point diffractions have known responses and by using time migration or depth migration these can be collapsed to the scatter point. The use of numerous points in modeling along the boundaries to act as absorber or sponges is the basis for a numerical boundary condition. This field of scatter points is regular and recent work shows clearly the effects of dense scattering points [14]. How this applies to radar clutter scattering is not yet known but the seismic response is dramatic for both acoustic and elastic wave equations. Further, we have developed a mathematical basis for the numerical method, creating a link to the previous analytic methods. Essentially, the sponge boundary condition has the features of the angle dependent analytic methods with a systematic scheme for specifying the angles.

In this report, we focus on reviewing a number of traditional seismic algorithms, especially the finite-difference reverse-time migration algorithm and the phase shift migration algorithm. We also perform preliminary experiments on synthetic seismic data using these algorithms and analyze their implications in radar signals processing. We further discuss their merits and shortcomings, and provide a plan for future research of this project.

2. Algorithms review

2.1 Finite-difference reverse time migration algorithm

This method is one of the depth migration methods that use finite difference solutions to the acoustic or elastic wave equation with the time reversal of the seismic traces being employed as time varying sources at the surface boundary. Migration is a technique to position seismic reflections in their correct subsurface location based on recorded seismic traces on the surface. A reflector can be treated as a large set of point-diffractors in the subsurface when seismic waves are refracted. This time reversal allows the wave field to be back propagated to the depth point where the reflections originated. Finite-difference reverse time migration algorithm is the most general method of all depth migration codes although it is also the most computer intensive. Using the complete wave equation allows energy to move in all directions during the back propagation process. The one way wave equation methods only allow propagation into the model from the boundary surface. If energy is "multiply reflected", bouncing back and forth between impedance surface changes the complete wave formulation allows for these effects.

The basis for this migration of the stacked seismic time section is the exploding reflector model [1]. For zero-offset case (coincident source–receiver situation), the response for a wave field propagating up from the reflector at half-velocity of the medium is considered equivalent to the a wave fields propagating downwards from the source to the reflector and then back to the receiver at the medium velocity. So the exploding reflector model assumes that if the downward wave travels the same path as the reflected upward wave, then only one wave propagation is needed. The upward wave will be chosen for modeling because it is the one that is recorded at the surface, and the migration will attempt to reconstruct the depth image of the exploding reflector wave field from the knowledge of the surface recorded wave fields.

A migration problem can then be viewed as an attempt to use data recorded at earth surface $z=0$ in order to reconstruct the depth image of the exploding reflector at $t=0$. In other words, migration attempts to reconstruct the exploding reflector wave field $u(x,y,z,t=0)$ from the knowledge of earth surface recorded wave field u(x,y,z=0,t) [3]. Good results with reverse time migration depend on general algorithm, good input data, and an accurate velocity model.

<u>(a) Forward modeling (for data generation)</u>

The zero offset data satisfy the wave equation

$$\frac{\partial^2 \Psi}{\partial t^2} = c^2(x,z)\left[\frac{\partial^2 \Psi}{\partial x^2} + \frac{\partial^2 \Psi}{\partial z^2}\right] \tag{1}$$

where *c(x, z)* is the velocity function (acoustic sound speed)
$\Psi(x,z,t)$ is the pressure wave fields
(x,z,t) are surface, depth, and time coordinates.

This equation provides a relationship between the second temporal derivatives and the second spatial derivatives. The finite difference approximation of the wave equation (1) is

$$\frac{\partial^2 \Psi}{\partial t^2} = \frac{\Psi^{N+1} - 2\Psi^{N} + \Psi^{N-1}}{\delta t^2} \tag{2}$$

Solving for the new time *N+1,* we get

$$\Psi^{N+1} = 2\Psi^{N} - \Psi^{N-1} + \delta t^2 c^2 \left[\frac{\partial^2 \Psi}{\partial x^2} + \frac{\partial^2 \Psi}{\partial z^2}\right]_N + src(t) \tag{3}$$

where *src (t)* is a source function which helps in generating artificial seismic pressure waves and in recording their reflections on the surface. The result from this step will be the synthetic data (time section). An example source function is given below which is plotted in Figure 1.

$$src(t) = \sin(2\pi f_{\max} t)\frac{e^{(-\alpha t^2)}}{\beta}$$

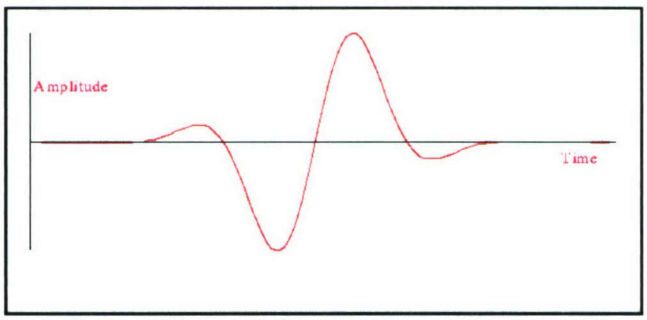

Figure 1 Source Function

(b) Reverse Time Imaging

The seismic data are introduced to the wave equation as boundary condition (data would propagate into earth from the subsurface receiver position), and the wave equation is then formulated as a backward in time propagation until time reaches zero (imaging condition).

Using the exact same wave equation but now solving for Ψ^{N-1}, we get

$$\Psi^{N-1} = 2\Psi^{N} - \Psi^{N+1} + \delta t^2 c^2 \left[\frac{\partial^2 \Psi}{\partial x^2} + \frac{\partial^2 \Psi}{\partial z^2} \right]_N + B.C_N \tag{4}$$

2.2 Phase shift migration algorithm

The phase shift method of migration is based on the numerical solution of the wave equation in the frequency domain with the initial conditions defined by the zero offset seismic section. The zero offset data are the wave fields {Ψ(x, z=0, t)} measured at some specific depth from the surface of the earth. This method begins with the Fourier transform of the data set. This transformed data is then extrapolated downwards and subsequently evaluated at t=0. This method is a one way transform of the wave equation.

In case of horizontal layered velocity, migration is defined by a set of independent ordinary differential equations in the *k-w* domain, and the wave components are extrapolated downwards by rotating their phases [7]. In case of lateral velocity variations, the wave field is extrapolated by phase shift methods using **n** laterally uniform velocities. The intermediate result is **n** reference wave fields. Then the actual wave field is computed via interpolation from the reference wave fields [8].

(a) Horizontal Velocity Variation

The pressure wave fields *Ψ(x, z=0, t)* (zero offset data) recorded at the surface satisfy the wave equation

$$\frac{\partial^2 \Psi}{\partial z^2} = \frac{1}{v^2}\left[\frac{\partial^2 \Psi}{\partial t^2}\right] - \frac{\partial^2 \Psi}{\partial x^2} \tag{5}$$

The process of migration is reconstructing, at the reflectors, the signals that produce the seismic section when propagated to the surface, in other words trying to evaluate *Ψ(x, z, t=0)*. These values are obtained by solving equation (5) with initial values *Ψ(x, z=0, t)*.

To solve for *Ψ(x, z, t=0)*, let *Ψ(x, z, t)* be represented by Fourier series

$$\Psi(x,z,t) = \sum_{k_x}\sum_{\omega} \Psi(k_x, z, \omega)\exp\left[i\left(k_x x + \omega t\right)\right] \tag{6}$$

where k_x is the midpoint wave number, and ω is the frequency.

Substituting (6) to (5) gives

$$\frac{\partial^2 \Psi}{\partial z^2} = -k_z \Psi \qquad (7)$$

The analytical solution to equation (7) is

$$\Psi(k_x, z+\Delta z, \omega) = \Psi(k_x, z, \omega) \exp\left[ik_z \Delta z\right] \qquad (8)$$

where $k_z = \pm \frac{w}{v}\left[1-\left(\frac{k_x v}{w}\right)^2\right]^{1/2}$

It should be noted that the two values of k_z correspond to forward time (negative value) and reverse time (positive value) wave propagation. Downward extrapolation requires that Δz be greater than zero. Since downward extrapolation of the recorded seismic data is an inverse problem, only the positive value of k_z is used in the solution.

The final expression for the wave extrapolation is thus given by

$$\Psi(k_x, z+\Delta z, w) = \Psi(k_x, z, w) \exp\left\{ i\frac{w}{v}\left[1-\left(\frac{k_x v}{w}\right)^2\right]^{1/2} \Delta z \right\} \qquad (9)$$

This is the solution to the wave equation

$$\frac{\partial \Psi}{\partial z} = i\left(\frac{\omega}{v}\right)\left[1-\left(\frac{k_x v}{\omega}\right)^2\right]^{1/2} \Psi \qquad (10)$$

Equation (9) is considered the basis of the phase shift method, and equation (10) is the exact extrapolation equation for the case of depth velocity variation with v=v (z).

Finally, substituting equation (9) into the Fourier series (6) and evaluating at t=0 will provide the solution to the migration problem for the depth variable velocity case.

$$\Psi(x, z+\Delta z, t=0) = \sum_{k_x}\sum_{\omega} \Psi(k_x, z, \omega) \exp\left[i\left(k_x x + k_z \Delta z\right)\right] \qquad (11)$$

Equation (6) can also be written as

$$\Psi(x, z, t=0) = \sum_{k_x}\sum_{\omega} \Psi(k_x, z=0, \omega) \exp\left[i\left(k_x x + k_z z\right)\right] \qquad (12)$$

where $\Psi(x, z=0, t)$, the coefficient of the series, represents the Fourier transform of the seismic section.

<u>(b) Lateral Velocity Variations</u>

In this case the solution expressed by equation (9) is no longer valid for the fields with lateral velocity variations, and the square root of equation (10) is expanded into truncated series (i.e. Taylor series).

3. Preliminary experiments

We have performed some preliminary experiments using the code developed for reverse time migration algorithm and the phase shift migration algorithm as provided in the open source Seismic Unix package.

3.1 Reverse time migration algorithm

This algorithm is illustrated below with synthetic data. A layered salt synthetic model is shown in Figure 2 and the corresponding migrated image using reverse time migration algorithm is shown in Figure 3.

Surface coordinate x

Depth (z)

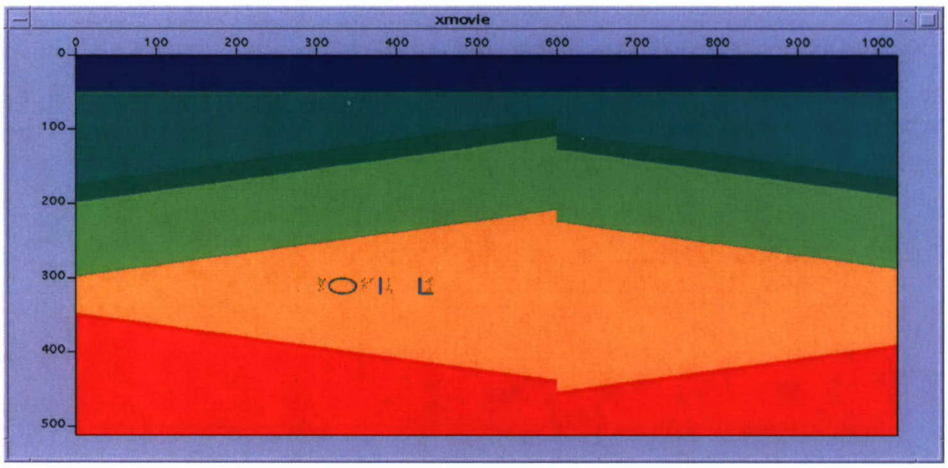

Figure 2 A salt model for synthetic data generation

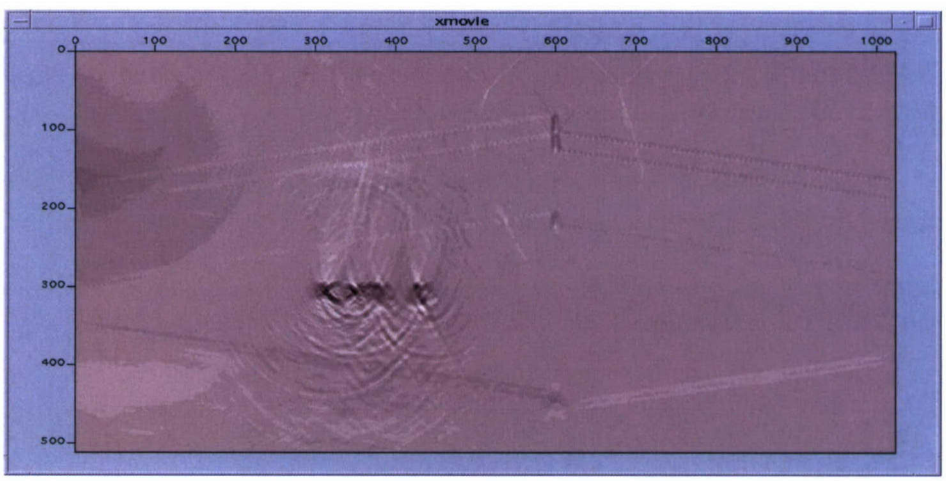

Figure 3 Migrated image for model in Figure 2

Figure 4 shows the input velocity model which consists of three reflector segments, a horizontal reflector and the other two reflectors with dips 45 and 60 degrees. The velocity model is a simple constant model with v=1500 m/sec, and the model dimension is 400x200 with grid spacing $\Delta x = 25$, $\Delta z = 25$.

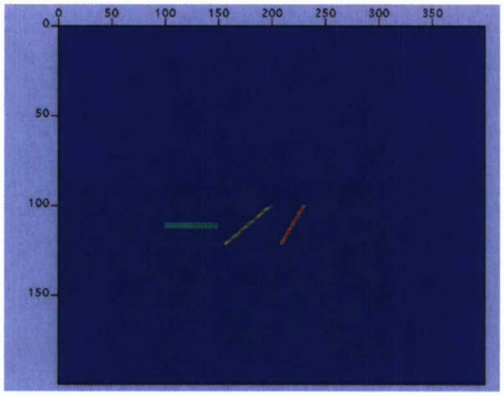

Figure 4 Input velocity model

The time section or the data resulting from the forward modeling, and the migrated image for the model are shown in Figures 5 and 6, respectively.

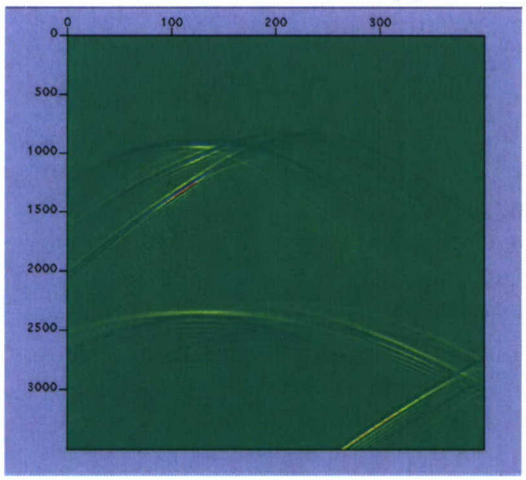

Figure 5 Zero-offset seismic section on the input model

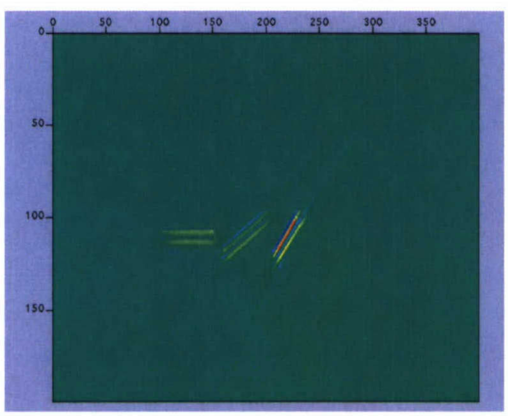

Figure6 The reverse time migrated result

As can be seen from the above figures it is noticed that the dips and the location of the reflectors and layers agree with those in the original models.

4. Discussion

The scattering of waves by clutter causes noise-like effects mixed with coherent signal. Because of the multiple like processes which are familiar in geophysical processing of seismic data we believe that using the complete wave equation should be our first attempt at modeling and migration.

An exploding reflector and reverse time migration (RTM) code was developed for the 2D acoustic wave equation. This code allows for a heterogeneous velocity model, a general gridded function of X and Z. Geophysics uses X for the horizontal axis and Z for the depth, positive downward. Our first experiment is to use a blocky earth model to generate a zero offset (X,T) data set. Then using the reverse time migration code as described above we generated the migrated image of the original model. Of course if the parameters are known exactly this should be an excellent recovery of the model as we demonstrate. The next experiment is to use a constant velocity background and a set of finite reflectors placed at different angles. These are then used as the source locations for the exploding reflector modeling phase. The RTM of this data gives an excellent result.

We plan on creating point diffractors in the region around these finite reflectors to simulate the effect of clutter. Currently, we are investigating small variations of the velocity model around a central mean value. These velocity gradients are intended to simulate the clutter velocity shift, the reduction in velocity due to scattering. These gradients are seen in ground penetrating radar (GPR) data and significantly better images are obtained in the GPR images when small changes in velocity are considered.

The phase shift methods for migration are suitable for velocity models which vary in only one dimension. These v(z) models are simple to implement in the method as each step downward requires a *z* increment. These methods are being tested with the same exploding reflector zero-offset data that is being used for the RTM process.

5. Conclusions

This is a technical report for the project "A Study of Inverse Methods for Processing of Radar Data" supported by the Air Force Research Lab. This project started on July 1, 2004 and completed August 31, 2005. The goal of this project is to investigate the possibility of new inversion algorithms for radar image processing to improve signal quality and reduce the effects of clutter based on study of known geophysical inversion algorithms. This technical report summarized what has been done concerning algorithm review and testing during this period of time.

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